



Invited review

Impact of Arctic Oscillation on the East Asian climate: A review

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ABSTRACT

The Arctic Oscillation (AO), which depicts a most dominant large-scale seesaw between the mid-latitudes and Arctic atmospheric mass, influences climate over Eurasia, North America, eastern Canada, North Africa, and the Middle East, especially during boreal winter. This review, with a special focus on the East Asian region, summarizes the climatic impact of AO. It begins with a description of the spatial structure of AO and the related climatic anomalies. The relationship of winter AO with the simultaneous East Asian winter climate (e.g. the East Asian winter monsoon (EAWM), cold surges/cold waves, and precipitation) and its instability are then followed. It is generally accepted that, through impacting the Siberian high, westerly wind, blocking frequency, Rossby wave activities etc., a positive phase of winter AO is associated with a weaker-than-normal EAWM, warmer conditions in East Asia, less frequency of cold surges/cold waves, increasing (decreasing) of winter precipitation in south (north) parts of East Asia; and vice versa. Notably, the pathways that the winter AO exerts impact are different. Besides, the AO-EAWM and the AO-cold surges/cold wave linkages have spatial and temporal variations. Subsequently, an overview of the inter-seasonal linkages between the East Asian summer monsoon with the preceding spring/winter AO is presented. There is a generally accepted knowledge that a positive spring AO is followed by significant positive summer precipitation anomalies in southern China and western Pacific as well as negative ones in the lower valley of Yangtze River and southern Japan. Finally, this review synthesizes the impact of winter/spring AO on the East Asian spring climate (e.g. dust storm, temperature, and precipitation) and discusses the potential predictive value of AO. The projection of AO and its impact on the East Asian climate in future has been barely explored. We conclude that, along with the long-term observation data, the linkage between AO and the East Asian climate on the sub-seasonal and decadal time scales, how tropical and extratropical forcing modulates the linkage and how the linkage evolves under future warming conditions should be more investigated. Notably, the change of AO during 1990–2013 winters could explain the Eurasian cooling but failed to explain the Arctic warming. In the future, the effect of Ural blocking on Arctic and Eurasian climate and their connection might be a hot topic.

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1. Introduction

Thompson and Wallace (1998) applied the empirical orthogonal function (EOF) to the extended-winter (November–April) sea level pressure (SLP) anomaly field over the domain poleward of 20°N and found that the leading vector of EOF is characterized by opposite pressure anomalies in the Arctic and the mid-latitudes (Fig. 1a). Although the AO resembles the North Atlantic Oscillation (NAO, Rogers, 1981) in many respects, it shows some unique characters in the Polar region and North Pacific. The main differences are 1) the NAO has no center of action in the Pacific, and 2) the AO has a broader center of action over the polar cap, giving it a more zonally symmetric appearance (Thompson et al., 2000). Furthermore, this leading EOF is more strongly coupled to surface air temperature (SAT) fluctuations over the Eurasian continent than the NAO (Thompson and Wallace, 1998; Wang et al., 2005). The correlation coefficients of this leading EOF time series and NAO index with the Eurasian extended-winter SAT anomalies during 1900–1995 are 0.55, 0.23 on the interannual time scale, respectively (Table 2 in Thompson and Wallace, 1998). To distinguish this pattern from the regional NAO, Thompson and Wallace (1998) referred to it as the Arctic Oscillation (AO) and the associated principal component time series as the AO index (Fig. 1b). Since then, the AO has been well-known in scientific community for its role in shaping the climate over the Northern Hemisphere (NH) (Gong et al., 2001; Gong and Ho, 2003; Jeong and Ho, 2005; Shi and Bueh, 2011; Kim and Ahn, 2012; Cui et al., 2013; Xue et al., 2014; Wan and Li, 2015).

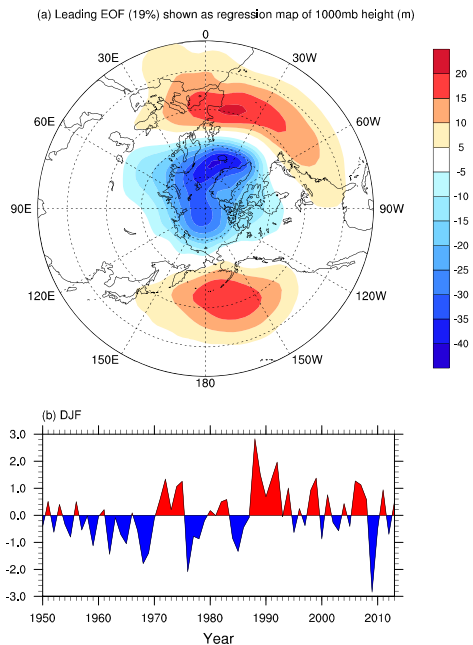


Fig. 1. The AO pattern. (a) Leading pattern of the AO which is defined as the leading mode of EOF analysis of monthly mean 1000 mb height during 1979–2000 period. (b) The principal component time series of the leading EOF during 1950–2013 winter (DJF). Figures and datasets could be downloaded from NOAA Climate Prediction Center on their website at http://www.cpc.ncep.noaa.gov/products/precip/CWlink/daily_ao_index/ao.shtml.

The AO is present in both cold and warm seasons, but its amplitude and meridional scale are generally larger during the cold season (Thompson and Wallace, 2000). It is revealed that the positive polarity of winter AO is associated with positive SAT anomalies throughout high latitudes of Eurasia. This SAT anomalies pattern is evident throughout the year except during the boreal summer months (Thompson and Wallace, 2000). Due to the close interannual relationship between the winter AO and Eurasian winter SAT, some studies suggest that the transition of AO from negative phase to its positive phase after the mid-1980s might contribute to interdecadal warming over Eurasia. For example, Thompson and Wallace (2000) pointed out that during 1970–2000 January–March, positive trends in the AO described ~50% of the warming trends over Europe and Asia, ~30% of the warming over the entire Northern Hemisphere, and ~40% of stratospheric cooling. Meanwhile, many previous studies have documented that the climate over Europe and West Asia is also largely dominated by the NAO (Hurrell and Deser, 2009; Luo and Cha, 2012; Luo et al., 2014, 2016a, 2016b). It's very difficult to distinguish the relative effects of NAO and AO over these regions. And it is not the ambition of this review paper to discuss the difference between AO and NAO because it is still a debatable question (Deser, 2000; Ambaum et al., 2001; Wallace and Thompson, 2002). Besides, the various alterations of meteorological parameters over the Atlantic Ocean and Europe associated with NAO have been sufficiently reviewed by Bader et al. (2011). Therefore, the impact of AO over East Asia will be the main focus of this review paper.

East Asia, located in the eastern Asian continent, includes the countries such as China, North Korea, South Korea, Mongolia, and Japan (please see Fig. 6a for geographical position). Driven by the thermal contrast (both in winter and summer) between the Eurasia continent and the Pacific, East Asian monsoons are the dominant climate over East Asia. The East Asian winter monsoon (EAWM) (generally refer to December–January–February) consists of sub-components, such as the Siberian high, Aleutian low, East Asian trough, low-level northerly and high-level East Asian jet stream (Wang and He, 2012; He and Wang, 2013a; He et al., 2013; He, 2015b). The low-level northerly along coastal East Asia brings cold air to the Asian continent, leading to cold climate during winters (Boyle and Chen, 1987; Chang et al., 2006; Li and Wang, 2012b; Li and Wang, 2012a, 2013). Another distinctive component of the Asian climate system is the East Asian summer (generally refer to June–July–August) monsoon (EASM). As the anomalous southerly (lower component of the EASM) are expected to bring more moisture to East Asia, the summer precipitation over East Asia is significant correlated with the EASM (Wang, 2001, 2002). In fact, the EASM can contribute to about 40–50% of the annual total precipitation amount over East Asia (Gong and Wang, 2003).

Given the significant influence of monsoon on East Asian climate, there has been much research effort devoted to better understanding the ingredients of the cause of the East Asian climate variability which shows both the interannual and interdecadal change. The EAWM varies on interannual timescales with strong variability on quasi-2, 4, and 4–7 years and on interdecadal timescales with remarkable cycles of 9–10, 20–30, 40 years and even longer (Ding et al., 2014, 2015). Early studies suggested that the East Asian climate variability could be significantly impacted by the anomalies of global ocean-land-atmospheric interaction such as El Niño–Southern Oscillation (ENSO) (Wang et al., 2000; Chang et al., 2006; He, 2015a), snow cover over the Eurasian continent (Shinoda, 2001; Liu and Yanai, 2002), AO (Gong et al., 2001; Jeong and Ho, 2005; Park et al., 2010), NAO (Luo et al., 2016a, 2016b),

and Arctic sea ice (Jhun and Lee, 2004) on the interannual timescale as well as the Pacific Decadal Oscillation (Kim et al., 2014), the Atlantic Multi-decadal Oscillation (Li and Bates, 2007; He and Wang, 2013a), solar activity (Zhou et al., 2007), and anthropogenic aerosols (Lau and Kim, 2006) on the interdecadal timescale. Among these factors affecting the East Asian climate, the studies of AO started relatively late but sprung up for more than decade. Through influencing the large-scale circulation such as East Asian jet stream, Siberian high, stationary planetary wave, a positive (negative) phase of winter AO generally leads to a weaker (stronger) EAWM and warmer (colder) than normal winters (Gong et al., 2001; Wu and Wang, 2002a; Wu and Wang, 2002b; Jeong and Ho, 2005; Park et al., 2010; He, 2013). These air temperature anomalies are mainly located in Northeast Asia. Meanwhile, many previous studies have noted that the EASM shows an evident linkage with the preceding winter/spring AO. For example, a significant out-of-phase correlation between the spring AO index and the summer precipitation along the Meiyu-Changma-Baiu rain-belt has been found (Gong and Ho, 2003; Gong et al., 2011). On the other hand, many other studies revealed that the winter/spring AO could impact the East Asian spring climate including the dust storm, temperature, and precipitation (Liu and Ding, 2007; Mao et al., 2011b).

In this review, we will summarize the published literatures on the impact of the AO on the East Asian climate, spanning from spring, summer, and winter. More specifically this review will discuss:

- the relationship of the winter AO with the EAWM, cold surges/cold waves, and winter precipitation over East Asia and its spatial and temporal variability;
- the impact of the winter/spring AO on the EASM precipitation and the instability of the relationship between the spring AO and the EASM on interannual and decadal time scales;
- the influence of winter/spring AO on the variability of spring dust storm, air temperature, and precipitation over East Asia;
- the potential predictability of the East Asian climate using the preceding AO signals.
- a summary of mechanisms that AO exerts impact is given in each part for general readers, though different pathways have been suggested and no consensus has been reached.

2. East Asian winter climate and winter AO

East Asian continent is much colder than most regions at similar latitudes in winter. Thus, cold surges (during which the Siberian high is extremely stronger and the day-to-day drop SAT exceeds at least 1.5σ (σ is the standard deviation of SAT)) are the most profound climate feature in winter, impacting China, Korea, Japan, and surrounding regions. One of the main factors inducing cold winters to East Asia is the EAWM. Interestingly, the winter AO could impact the East Asian winter climate directly or through its impact on the EAWM.

2.1. East Asian winter monsoon and winter AO

It is generally accepted that a weaker-than-normal EAWM is associated with positive phase of AO and vice versa. Gong et al. (2001) is among the first to investigate impact of AO on the EAWM. Based on the monthly mean air temperature data of 160 stations in China and NCEP/NCAR (National Center for Environmental Prediction/National Center for Atmospheric Research) reanalysis dataset, Gong et al. (2001) found a significant out-of-phase relationship between the AO and the EAWM during 1951–1999. That is, negative phase of AO is concurrent with a stronger East Asian trough and an anomalous anticyclonic flow over Urals at the middle troposphere (500 hPa). They suggested that the AO could significantly influence the EAWM through its impact on the Siberian high. On the basis of 40 years (1958–1997) daily-mean data from the NCEP/NCAR Reanalysis, Thompson and Wallace (2001)

revealed that cold events (in which daily minimum temperature drops below a specified threshold corresponding to 1.5σ below the local seasonal mean) in late winter (January to March) occur with much greater frequency over East Asia under low AO index conditions. Meanwhile, low AO index conditions are marked by decreased 500-hPa height variance across the North Pacific stormtrack from East Asia into the Northeast Pacific, which means that the East Asian trough is deepened and the EAMW is stronger. A pronounced trend toward AO high-index polarity since the late 1960s has favoured warmer winters across most of the North Hemisphere high-latitude continents and a decreased incidence of high-latitude blocking (Thompson and Wallace, 2001). Using the NCEP/NCAR reanalysis dataset spanning from 1958 to 1997, Wu and Wang (2002a) investigated the connection between the winter AO and the Siberian high, and the EAWM. Their results suggested that when the winter AO is in its positive phase, significant anomalous upward motion appears in nearly entire troposphere from 60°E to 80°E at mid-latitudes (around 50°N). As the maintaining of winter Siberian high primarily depends on downward motion of airflow aloft (Ding et al., 1991), that is, the largest downward motions anomalies appear in mid-latitudes from 50°E to 90°E in associated with stronger winter Siberian high. Correspondingly, the Siberian high is remarkably weaker than normal during positive AO phase, which further weakens the EAWM. Thus, the winter air temperature is higher in Siberia, Mongolia, eastern China, Korea, Japan and the partly western North Pacific. Meanwhile, the AO could influence the cold polar air intrusions over East Asia through the associated wave activity (Wen et al., 2013). Chen et al. (2005) found that a larger amount of wave activity flux propagates upward from the lower troposphere at mid-latitudes such that the polar vortex gets warmer and weaker during negative AO phase, leading to the weaker westerlies circulating the polar region which favor the southward advection of polar air and lead to a cooler Eurasia.

Some studies suggested that the AO could impact the EAWM without the effect on the Siberian high. Using the NCEP/NCAR Reanalysis, Wu and Wang (2002b) suggested that the winter AO influences directly the SAT, SLP, and the East Asian trough at 500 hPa over the regions northward of 35°N in East Asia rather than through its impact on the Siberian high. Such a conclusion is different from that of Gong et al. (2001) and Wu and Wang (2002a) but the evidence for supporting the direct impact of winter AO on the EAWM is not well addressed. Thompson and Wallace (2000) pointed out that the positive anomalies of winter SAT in Eurasia might be caused by the temperature advection induced by the enhanced westerlies associated with the positive phase of AO, which could bring more warm marine air into the interior of the continents. Similar results are obtained by Suo et al. (2009), who suggested that the temperature advection induced by both the anomalous zonal and meridional winds associated with AO is responsible for the temperature anomalies. On the other hand, the key role played by Siberian high in the influence of winter AO on the EAWM or winter SAT in East Asia was further investigated by Gong and Wang (2003). Gong and Wang (2003) demonstrated that the correlation coefficient between AO and average temperature in China during 1958/59–1994/95 winters is 0.43, which decreases dramatically to 0.14 when the contribution of the Siberian high is excluded. What's more, the interannual variability of the seasonal cycle of the EAWM could be explained to some degree by the AO. The EAWM tends to exhibit a stronger seasonal cycle (defined as the amplitude of seasonal change in a year) during negative phases of the AO and vice versa (Kim et al., 2013). Considering the negative relationship between the AO and the EAWM, some studies suggested that the weakening of the EAWM in the mid-1980s is attributed to the positive trend of AO (He, 2013; Lee et al., 2013).

2.2. Cold surges and AO

Cold surges exert tremendous social and economic impacts on East Asia (Park et al., 2011; Yu et al., 2015a, 2015b). Due to the significant

impact of AO on the EAWM, the linkage between the winter AO and the East Asian winter cold surges/cold waves (a rapid fall in temperature within a 24-hour period greater than 8 °C and the minimum temperature is less than 4 °C) has also drawn much attention. Generally, negative phase of AO leads to more frequent cold air outbreaks over East Asia (Thompson and Wallace, 2001; Wen et al., 2009; Park et al., 2010; Yu et al., 2015a). Jeong and Ho (2005) found that, during winters with negative AO phase, the SLP over the climatological position of the Siberian high (40°–60°N, 80°–120°E) is strengthened. When the Siberian high reaches a certain intensity, eastward moving upper level short-wave trough aloft over Lake Baikal deepens as it propagates toward quasi-stationary East Asian coastal trough. During the above process, surface anticyclone moves southward together with very cold air accumulated over eastern Siberia, which is called as a cold surge occurrence. Based on the daily mean surface temperature from 280 stations across northern China and European Centre for Medium-Range Weather Forecasts (ECMWF) 40-year Re-analysis datasets (ERA-40), Wei and Lin (2009) revealed a significant negative correlation between extended-winter cold wave frequency in northern China and the simultaneous AO during 1960–2001. They further suggested that the atmospheric circulation anomalies in Siberia might serve as a bridge for interaction between AO and cold wave frequency in northern China on inter-annual time scale. Many other studies revealed that the exceptionally negative AO in December 2009 contributed to the coldest weather in Eurasia (Cohen et al., 2010; L'Heureux et al., 2010; Wang and Chen, 2010). Wen et al. (2013) revealed that the strongest negative polarity AO, which persisted from December 2009 to March 2010, caused the extremely cold weathers in winter 2009–2010 over the Northern-Hemisphere high-latitude continents. Their analysis indicated that the stratospheric polar night jet in winter 2009–2010 was generally weaker compared with the climatological mean, which benefits the upward propagation of planetary waves, causing a sudden warming in stratosphere. Then the polar vortex completely split and the polar region was dominated by positive temperatures and geopotential heights anomalies. Subsequently, this major warming event induced the strong negative stratospheric AO, which further propagated downward to troposphere and caused the extreme cold weather. Besides, AO is closely linked to severe cold surges in East Asia through modulating the formation of Ural-Siberian blocking highs. Associated with an intensified Ural-Siberian blocking high, the enhanced northerly cold advection increases mass convergence in the upper troposphere, which in turn intensifies the surface Siberian high and triggers a cold air outbreak in East Asia (Ding et al., 1991). This provides a fundamental linkage between the AO and the East Asian cold surge (Cheung et al., 2012).

The impact of AO on East Asian winter climate changes is spatially and temporarily dependent. An interesting phenomenon noticed by Yang and Li (2008) is that, even though the AO in January 2008 was in its positive phase, the Ural blocking high and EAWM were stronger than normal and South China suffered lasting extreme cold weather, which are opposite to their general situations related to positive AO. Early studies suggested that the EAWM is weaker (Gong et al., 2001) and East Asian winter temperature is higher than normal along with positive AO (e.g. Thompson and Wallace, 2001; Quadrelli and Wallace, 2002). However, a closer examination reveals that the influence of winter AO on air temperature is limited to high latitudes of Eurasia (north of 40°N) (see Fig. 4 in Quadrelli and Wallace, 2002). On the other hand, we do see positive AO is accompanied by significant decreased temperature over the central and southern China by inspecting the linear correlation between AO index and surface temperature during 1979–2008 January (see Fig. 12a in Wen et al., 2009). The paradox might be caused by the characteristic impact of AO in sub-seasonal time scale and the mismatch between the stations observational data and the reanalysis dataset. Cold surges in East Asia could happen both in negative and positive phase of AO, but with distinct properties (Park et al., 2010). The cold surges during a negative AO were modulated by southward expansion of the Siberian high, which originated in the Arctic regions. On the other hand, the

cold surge during a positive AO occurred when the Siberian high extended southeastward across the Eurasian continent. Park et al. (2011) further pointed out that the cold surges occurred in November–March over East Asia could be grouped into two general types: wave train and blocking types. The blocking type of cold surge tends to occur during negative AO periods. However, the wave train type is observed during both positive and negative AO periods, but the wave train associated with negative AO is relatively weaker. The cold surges during negative AO are stronger than those during positive AO in terms of both intensity and duration. Result revealed by Park et al. (2010, 2011) suggested that different AO phases are accompanied by distinctive characteristics of cold surges in terms of intensity, frequency, duration, and tracks.

Using measured air temperature from 542 meteorological stations in China, Chen et al. (2013a) demonstrated a significant negative correlation between the AO and the number of extreme cold days (the daily minimum temperature was lower than the 10th percentile of the daily minimum temperature during the period 1961–2011) in north (north of 40°N) and central (between 30°N and 40°N) East China (East of 105°E) during the period 1961–2011. However, the number of extreme warm days (the daily maximum temperature was higher than the 90th percentile of the daily maximum temperature during the period 1961–2011) across East China showed little correlation with AO during the same period. Kim and Ha (2014) identified two principal modes of winter blocking variability. One mode is characterized by regional blocking activities including the North Pacific, Greenland, European, and Ural-Siberian regions. The other mode shows a zonally dipole pattern between North Pacific and North Atlantic blockings. It is revealed that the phase of North Pacific Oscillation (NPO) and AO could controls the dominant modes of blocking variability. The negative NPO (AO) is favorable for the in-phase (out of phase) relationship between west of (south of) North Pacific and south of (west of) North Atlantic blocking frequency. Besides, when the NPO and AO are in phase (out of phase), there exists enhanced (reduced) Northern Hemisphere blocking occurrence, in particular, having higher frequency over the Pacific and higher (lower) frequency over the Northwest Atlantic. Similarly, the influence of the winter AO on the simultaneous temperature over East Asia might be different in different phase of Western Pacific (WP) teleconnection in the North Pacific (Park and Ahn, 2015). Corresponding to winter with negative (positive) AO concurrent with negative (positive) WP, the winter temperature shows robust dominant negative (positive) anomalies across the entire East Asia region. By contrast, the East Asian winter temperature barely shows significant anomaly when the AO and WP are out-of-phase. Lim and Kim (2016) revealed that the influence of winter AO on the interannual variability of EAWM and East Asian winter temperature during warm Eurasian teleconnection phase is significantly different from that during cold Eurasian teleconnection phase. It is suggested that the East Asian winter temperature shows significant warm (cold) anomalies when the warm (cold) AO coincides with warm (cold) Eurasian teleconnection phase.

The impact of winter AO on Eurasian SAT and cold surges might have decadal change (Woo et al., 2012). As revealed by many previous studies that during positive phase of AO, positive SAT anomalies elongated over Europe to East Asia, while strong cold anomalies appear over Greenland. However, the amplitude of positive SAT anomalies related to positive phase of AO is larger in 1990s and 2000s than 1980s (Woo et al., 2012). Besides, the negative SAT anomalies associated with negative AO phase in the 1980s and 2000s are remarkably different from those in the 1990s over the Eurasian continent. Negative SAT anomalies in the 1980s and 2000s elongate over Europe to East Asia, whereas negative SAT anomalies in the 1990s are mainly confined over Europe, Northern Siberia and Kazakhstan. The amplitudes of cold anomalies over Europe in the 1990s are also much weaker than those in the 1980s and 2000s. Meanwhile, the cold surge events in the 1980s (58.0%) and 2000s (54.7%) occur more frequently in the negative phase of the AO than in the positive phase, while in 1990s more cold

surge events (61.5%) appear in the positive phase of AO. Besides, mean duration (strength) of cold surges with negative AO phase is longer (stronger) than that with positive AO phase in all of the three decades, which might be attributed to the stronger Ural blocking anomaly and longer duration of East Asian trough in the negative AO phase. [Chen et al. \(2013a\)](#) pointed out that even though a significant connection existed between winter extreme cold days in north Eastern China and the AO, whether before or after the shift in the mid-1980s, a significant connection between winter extreme cold days in central Eastern China and the AO was only found for the period 1961–85. It implies that the influence of AO on winter temperature extremes is spatially and temporarily dependent, which might be attributed to the unstable relationship between the AO and the EAWM. As revealed by [Li et al. \(2014\)](#) that there is a robust strengthening of the EAWM-AO relationship after the early of 1980s ([Fig. 2](#)). To summarize:

- (1) The Siberia high might serve as a bridge for interaction between AO and cold wave frequency over East Asia (e.g. [Jeong and Ho, 2005](#); [Wei and Lin, 2009](#)). The cold surges during a negative AO were modulated by southward expansion of the Siberian high, which originated in the Arctic regions. On the other hand, the cold surge during a positive AO occurred when the Siberian high extended southeastward across the Eurasian continent (e.g. [Park et al., 2010](#)).
- (2) The anomalous stratospheric process (e.g. sudden warming) could induce anomalous stratospheric AO events, which further propagate downward to troposphere and cause the extreme cold weather over East Asia (e.g. [Wen et al., 2013](#)).
- (3) The large amount of wave activity flux associated with anomalous surface negative AO events propagates upward from the lower troposphere at mid-latitudes such that the polar vortex gets warmer and weaker, leading to the weaker westerlies circulating the polar region which favor the southward advection of polar air and lead to a cooler Eurasia (e.g. [Chen et al., 2005](#)).
- (4) The influence of AO on winter temperature extremes is temporarily dependent. For example, the negative SAT anomalies associated with negative AO phase in the 1980s and 2000s are remarkably different from those in the 1990s over the Eurasian continent (e.g. [Woo et al., 2012](#)). But the related reasons are still not clear.

2.3. AO, NAO, Ural blocking and warm Arctic–cold Eurasia

One of the most dominant feature of the climate change in the past century is the overall global warming. And the clearest manifestation of global warming is the enhanced warming over the Arctic where the SAT has risen twice as large as the global average in recent decades - a feature called Arctic amplification ([Bekryaev et al., 2010](#); [Gao et al., 2015](#)). Along with the Arctic amplification, a band of cooling trend

since the 1990s that extends across the mid-latitudes of Eurasia has been noticed by recent studies ([Outten and Esau, 2012](#); [Cohen et al., 2014](#); [Kug et al., 2015](#)). As pointed out by [Cohen et al. \(2014\)](#) that continental winter SAT trends since 1990 exhibit cooling over the mid-latitudes. The negative trends extend from Europe eastward to East Asia, with a center of maximum magnitude to the west of the Baikal ([Fig. 3a](#)). As reviewed above, the AO/NAO shows an in-phase relationship with the SAT over Eurasia. A natural question is whether the Eurasian winter cooling trend is related to the variability of AO or NAO? To illustrate this issue, [Fig. 3b](#) and [c](#) display the linear regression of winter SAT with respect to the negative AO and NAO indices, respectively. Obviously, when the AO is in its negative phase, winters are predominately cold across northern Eurasia. The regions from Europe to East Asia show significant negative SAT anomalies ([Fig. 3b](#)), the extent and negative center of which are well consistent with the spatial pattern of SAT linear trend ([Fig. 3a](#)). Meanwhile, the variability of winter SAT associated with simultaneous negative NAO also shows negative anomalies across northern Eurasia, though the significant anomalies are located more westward ([Fig. 3c](#)). Notably, the linear trend of winter AO/NAO during 1990–2013 is about $-1.49/-0.73$ per decade. Therefore, the negative trend in the AO/NAO might explain the recent Eurasian winter cooling ([Cohen et al., 2014](#)). However, the variability of

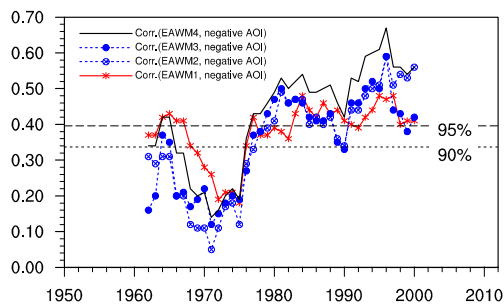


Fig. 2. The 25-yr sliding correlation coefficients between the four EAWM indices (EAWM1–4) and the negative AOI during 1950–2012. The results are obtained from NCEP/NCAR reanalysis datasets. Figure is repeated from [Li et al. \(2014\)](#).

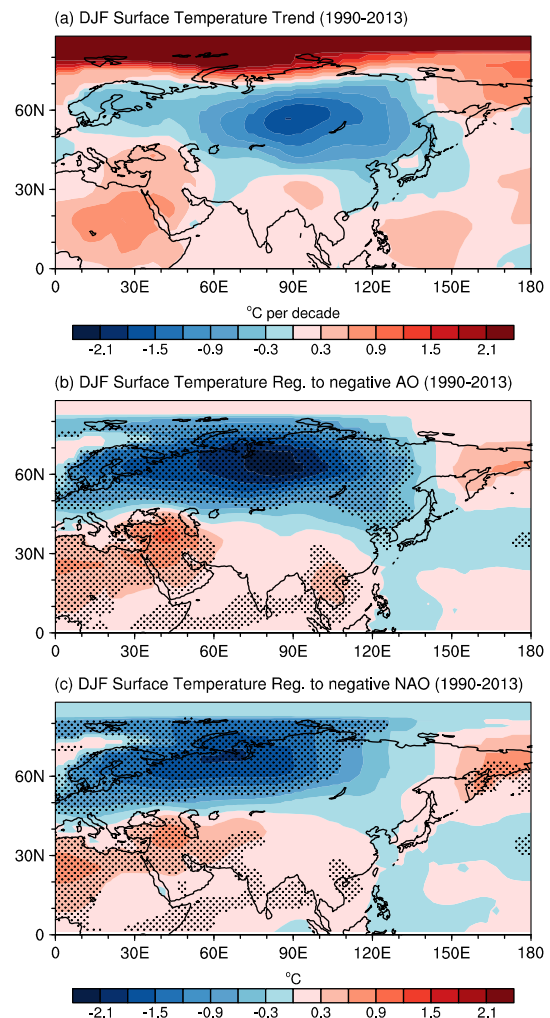


Fig. 3. (a) Linear trend ($^{\circ}\text{C}$ per 10 years) in winter SAT from 1990/91–2013/14. Figure is repeated from [Cohen et al. \(2014\)](#). Linear regression coefficients of winter mean SAT (units: $^{\circ}\text{C}$; shaded) upon the simultaneous negative AO (b) and NAO (c) index from 1990/91–2013/14. Stippled regions indicate the regression coefficients statistically significant at 90% confidence level, as estimated using the Student's *t*-test. SAT anomalies are derived from the NASA GISS temperature record ([GIStEMP Team, 2016](#)).

AO or NAO itself failed to explain the Arctic warming. Some recent studies suggested that the Arctic amplification might be related to the activity of the Ural blocking (Luo et al., 2016a, 2016b). Luo et al. (2016a) found that, along with the occurrence of strong Ural blocking, significant warming emerges on the northern side of the Ural blocking anticyclonic anomaly center with an obvious cold anomaly on its southeastern side, which resembles the warm Arctic-cold Eurasia pattern. It means that the Ural blocking could amplify the Arctic warming. More importantly, it is found that the effect of Ural blocking on the amplification of Arctic warming depends on the phase of NAO (Luo et al., 2016b). An obvious warm Arctic-cold Eurasia pattern is observed during the strong Ural blocking lifetime when the NAO is in its positive phase (Fig. 4a). By contrast, the Arctic warming is relatively weaker during Ural blocking events when the NAO is negative (Fig. 4b) or neutral (Fig. 4c) phase. It implies that the large scale atmospheric or climate modes such as AO and NAO might have some deficiencies in describing the regional climate variability. The Ural blocking might provide a new insight in exploring the mechanisms for the formation of warm Arctic-cold Eurasia, which should be paid more attention in further studies.

2.4. External forcing for the connection of winter AO with the EAWM

The connection between the winter AO and East Asian winter climate might be modulated by external forcing (such as solar cycle, SST, sea ice). The influence of positive/negative phase of AO on the East Asian winter climate occurring in El Niño winters differs from that in La Niña winters. Based on the NCEP/NCAR Reanalysis, Quadrelli and Wallace (2002) found that the structure of the AO in December–March is significantly different between warm and cold ENSO cycle. During warm ENSO winters, the Arctic center of action of the AO is more prominent and extends deeper into Siberia, thus the impact of the AO upon winter temperatures over much of Siberia and North East Asia is much stronger than that during cold ENSO winters. Similarly, Cheung et al. (2012) suggested that the November–March positive (negative) AO concurrent with El Niño (La Niña) could contribute to a stronger blocking-EAWM relationship. This is because the in-phase of AO and ENSO could reinforce the anomalous westerly in the mid-latitudes. In general, the positive phase of winter AO could lead to warming conditions over East Asia (Thompson and Wallace, 2000). Recent model simulation indicated that the emission of global black

carbon could induce positive phase AO responses (Wan and Li, 2015). On the other hand, the negative AO usually favors cold climate over Japan. However, exceptions are also present. For example, the negative AO persisted from October through December in 2012, however, Japan was relatively warm in October 2012 (Ando et al., 2015). Ando et al. (2015) suggested for the first time that the heating by warm waters in the oceans located in west of Japan could overwhelm the cooling effect of the strongly negative AO. It implies that the influence of winter AO might be modulated by other factors. Chen and Zhou (2012) suggested that the relationship between winter AO and East Asian winter climate might be modulated by the 11-year solar cycle. During winters with high solar activity (1959, 1960, 1967, 1968, 1969, 1970, 1971, 1979, 1980, 1981, 1982, 1983, 1989, 1990, 1991, 1992, 1993, 1999, 2000, 2001, 2002), robust SAT anomalies appeared in northern Asia associated with anomalous AO events. Meanwhile, significant anomalies in 850-hPa northerly wind over northeastern China and East Asian trough at 500 hPa were observed. However, during winters with low solar activity (1961, 1962, 1963, 1964, 1965, 1966, 1972, 1973, 1974, 1975, 1976, 1977, 1978, 1984, 1985, 1986, 1987, 1988, 1994, 1995, 1996, 1997, 1998, 2003, 2004, 2005, 2006, 2007, 2008, 2009, 2010), the anomalies of SAT, 850-hPa northerly wind and the East Asian trough were much weaker associated with anomalous AO events. They suggested that the AO tends to be more active during high solar activity winters than that during low solar activity winters, which exerts more significant influence on the climate of East Asia. Huth et al. (2007) also suggested that the intensity and annularity of AO depend on the 11-year solar cycle. The Pacific center of AO weakens and eventually disappears when the solar activity is moderate. Based on both observational datasets and model simulations, Li et al. (2014) suggested that the reduction of autumn Arctic sea ice cover could modulate EAWM-AO relationship. They found that the relationship of winter AO with the EAWM and East Asian winter SAT was statistically insignificant during 1950–1970 (Fig. 5a) whereas significant during 1983–2012 (Fig. 5b), which might be attributed to the East Asian jet stream (EAJS) upstream extension. Before the 1970s, the EAJS signal is relatively confined to the western North Pacific. After the 1980s, the near-surface heating over the Barents–Kara Seas in the eastern Arctic caused by the reduced sea ice is associated with a strong anticyclonic anomaly over the polar ocean and anomalous easterly advection over the northern continents. This causes a wider horizontal structure of eastward propagation of Rossby

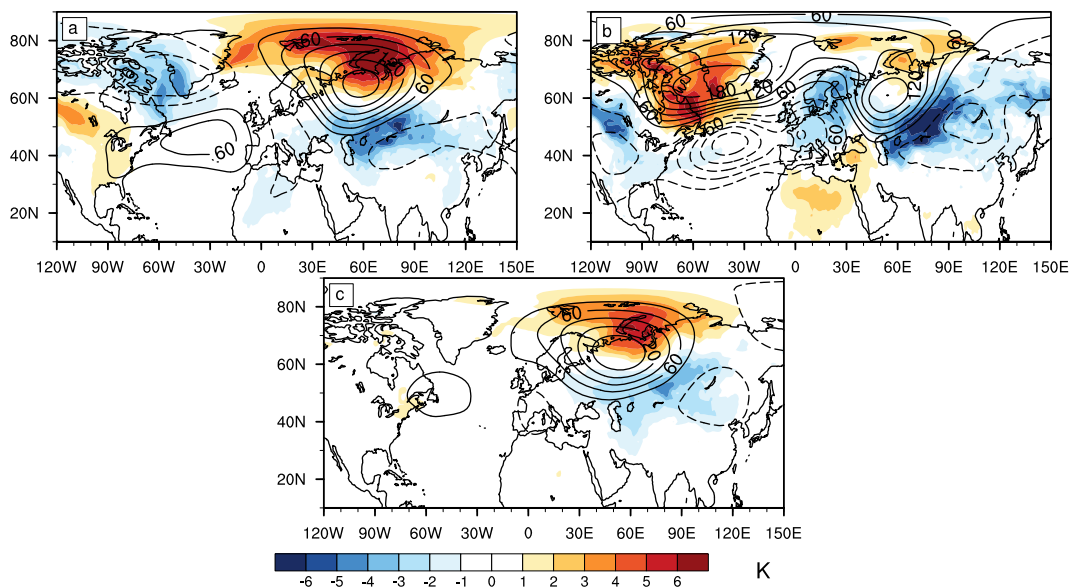


Fig. 4. Time-mean anomalies of 500-hPa geopotential height (contours, interval = 30 gpm) and SAT (colors; unit: K) for the Ural blocking events from lag –5 to 5 days for (a) positive NAO (13 cases), (b) negative NAO (5 cases) and (c) neutral NAO (36 cases) events for a 1.0 standard deviation case during winters of 1979–2013. Lag 0 denotes the day when the Ural blocking peaks. The contours and shaded values are above the 95% confidence level based on a two-sided Student's *t*-test. Figure is taken from Luo et al. (2016b).

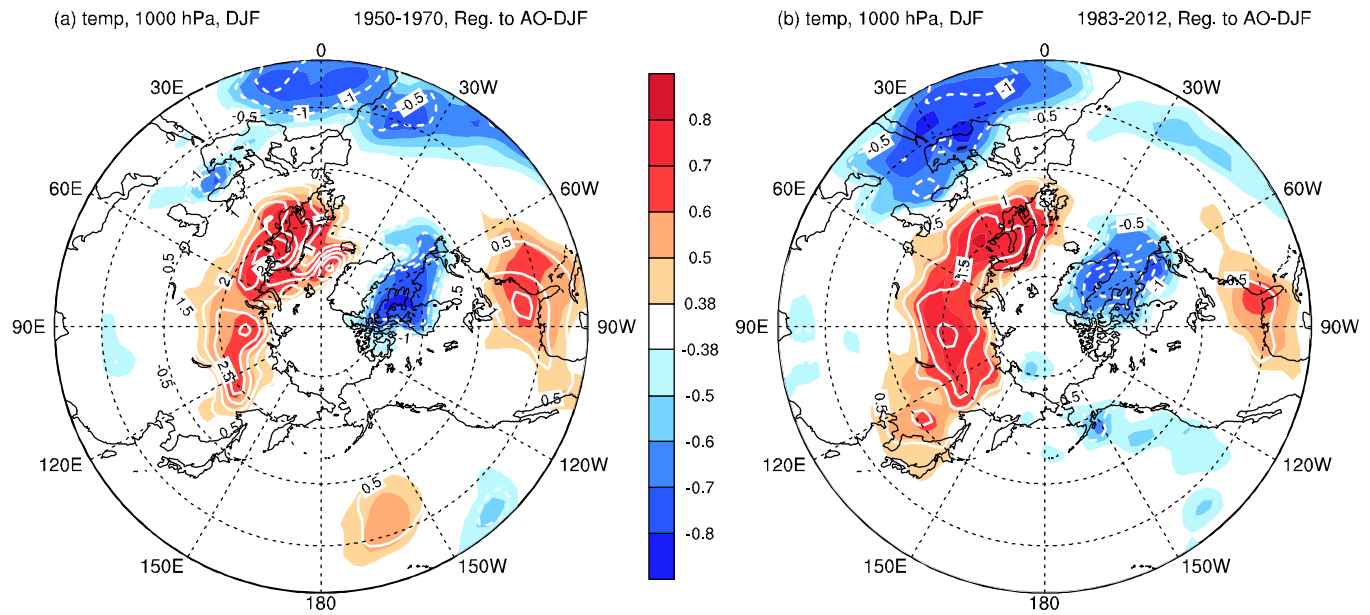


Fig. 5. Linear regression coefficients of winter 1000-hPa temperature (units: °C, contour) derived from NCEP/NCAR reanalysis dataset upon the simultaneous AO index during (a) 1950–1970 and (b) 1983–2012. Shaded values indicate the correlations that are statistically significant at 90% confidence level, as estimated using the Student's *t*-test. Figure is repeated from Li et al. (2014).

wave, which causes a considerable Eurasian cooling and further intensifies the meridional temperature gradient between the midlatitude and tropical regions. This favors the upstream extension of EAJs that bonds the EAWM and the AO. Such an unstable relationship is similar to the phenomena revealed by Zhao and Moore (2009). Zhao and Moore (2009) suggested that the winter AO is well coupled with the North Pacific SLP during 1925–1950 and 1980–1998 but decoupled during 1951–1979. They further pointed out that such a decadal change of AO's teleconnection with the North Pacific climate might be modulated by the Pacific Decadal Oscillation (PDO).

2.5. Winter AO and winter precipitation

Winters with positive AO are generally accompanied with positive precipitation anomalies in central and southwest China as well as negative ones in northwest and northeast China (Fig. 6). Due to enhancement of polar night jet stream, not only could positive winter AO lead to warm air advection from the North Atlantic to high latitudes of Europe and prevent cold air of polar region from invading high latitudes of East Asia, but also cause more precipitation (mainly snowfall) in the high latitude of Eurasia (Thompson and Wallace, 2001). Based on monthly precipitation data from 160 meteorological stations (mainly located in the east of China) in China and NCEP/NCAR Reanalysis data, Gong and Wang (2003) noticed that the positive winter AO is associated with positive precipitation anomalies in China, especially over the central China (30°N to 40°N, east of 100°E). The mean precipitation of the 160 meteorological stations correlates with AO at 0.47 during 1958/59–1994/95 winters (above the 95% confidence level). Interestingly, similar results could still be found during the extended period (1961–2013 winters) based on monthly mean precipitation derived from both the 2400 observing stations in China and NOAA's reconstruction dataset (Fig. 6). Park et al. (2011) suggested that the precipitation accumulated within 4 days after cold surge outbreak in November–March during positive AO is larger than that during negative AO over entire East Asia except west of Korea. They also suggested that the freezing precipitation or snow over inland China (Korea and Japan) is likely to occur more frequently during the positive (negative) AO periods.

Similar conclusion was drawn by Wen et al. (2009). On the basis of updated daily precipitation observations over 593 stations in China

and NCEP/NCAR reanalysis datasets, Wen et al. (2009) revealed a distinctive upper-tropospheric wave train extending from northeastern Atlantic, through the Mediterranean Sea and western Asia to southern Asia in associated with positive AO during 1979–2008 January. Meanwhile, the Middle East jet stream (MEJS, located over northern Saudi Arabia) intensifies and shifts southeastward. The intensification and southeastward shift of the MEJS are accompanied by cold and wet conditions over the majority of southern–central China. They pointed out that the intensified and southeastward shifted MEJS and positive AO

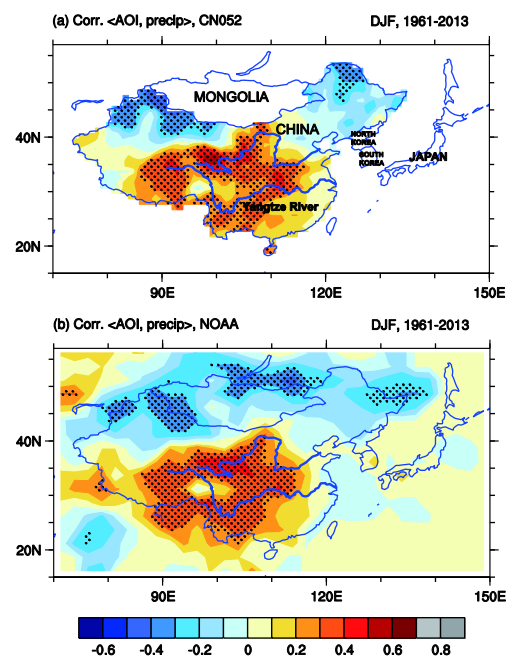


Fig. 6. Correlation coefficients between the winter AO index and simultaneous precipitation provided by Wu and Gao (2013) (a), which is based on the interpolation from 2400 observing stations in China during 1961–2014 and provided by NOAA (b). Dotted values are statistically significant at 95% confidence level, as estimated using the Student's *t*-test. Figure is drawn on the method of Gong and Wang (2003).

played important roles in the occurrence of the extraordinarily frequent and long-lasting snowstorms affecting China in January 2008. Yang and Li (2008) also revealed that associated with positive winter AO, there is more winter precipitation in most parts of China (except Mongolia and Xinjiang province) with a positive anomaly over coastal region of South China.

Based on the time-variable gravity data (equivalent water thickness deviation: EWD) from the Gravity Recovery and Climate Experiment (GRACE) satellite system and terrestrial water (including snow) storage given by the Global Land Data Assimilation System (GLDAS) Noah model, Matsuo and Heki (2012) discussed the precipitation anomalies caused by AO in the Northern Hemisphere during 2002–2012 January–March. The correlation coefficients between the wintertime EWD and AO index at each grid points in North Hemisphere indicated that positive (negative) correlations are located in high (middle) latitudes (reverses across the latitude 55°N in Eurasia continent and North America and 75°N). As to East Asia, there are positive correlations located in South China (about South of 30°N, 90°–120°E) and negative correlations centered around the Lake Baikal (see Fig. 5 in Matsuo and Heki, 2012). The spatial distribution of correlations in China between precipitation and AO is consistent with that revealed by Wen et al. (2009, see their Fig. 12b). Li and Wang (2012a) also found that the positive winter AO facilitates significant positive winter precipitation anomalies over South China, which is consistent with the results of Wen et al. (2009) and Matsuo and Heki (2012).

Some studies have also addressed the linkage between winter AO and precipitation extremes in East Asia. On the basis of daily precipitation dataset obtained from the China Meteorological Administration during 1954–2009 and NCEP/NCAR reanalysis, Mao et al. (2012) revealed a significantly positive correlation between AO and the frequency of extreme precipitation events over China (mainly located over central-southern China) during January to February. They suggested that a positive AO phase is accompanied by a stronger than normal MEJS and a deepened southern branch trough over the Bay of Bengal. The deepened southern branch trough enhances synoptic scale disturbances in vertical motions in the low to middle troposphere over central-southern China. More moisture transported by the deepened southern branch trough and active synoptic scale disturbances in vertical motion over central-southern China would provoke more extreme precipitation events there. The enhanced MEJS works as waveguide to induce a wave propagation, which originates in Europe and spreads across northern Africa to Northwest India, transporting the signal of AO to South China. Yang et al. (2012) suggested that the strongest negative phase AO might be an important reason for the extreme drought during 2009/2010 October–February over southwestern China, which is characterized with the lowest percentage rainfall anomaly and the longest non-rain days in the past 50 years. This anomalous AO is accompanied with the weakened MEJS, cyclonic anomaly over Arabian Sea, anticyclonic anomaly over Tibet and cyclonic anomaly over Lake Baikal. These circulation anomalies weaken the Southern Branch Trough which directly decreases the moisture transport toward the southwestern China. Meanwhile, the cyclonic anomaly over the Lake Baikal causes a deepened and westward shifted East Asian Major Trough so that dry cold air easily invades down to southwestern China. Additionally, the Arabian Sea cyclonic anomaly favors the westward extension of Western Pacific Subtropical High. By contrast, Northern China suffered a significant positive precipitation anomaly in 2009/2010 winter (Li et al., 2012), which, in many areas, was more than twice that of normal years and might be induced by the large-scale weather and climate anomalies in the Northern Hemisphere related to significant negative AO anomalies during the same period (Wang and Chen, 2010; Ripesi et al., 2012).

The AO is also relevant to the East Asian winter precipitation variability on inter-decadal scales. It is revealed that the winter climate in Southeast China (102.5°–122.5°E, 20°–35°N) experienced a trend from a dryer state in the 1970s via an increase during the 1980s toward

stabilization on wetter conditions commencing with the 1990s to 2010, in which the anomalous anticyclone over South Japan plays an important role (Zhang et al., 2014). They suggested that the increased precipitation in Southeast China after the 1980s is attributed to the interdecadal strengthening of winter AO.

3. East Asian summer climate and AO

The EASM, which is characterized by southerly wind over East Asia and the rainfall concentration in a nearly east-west elongated rainbelt, affects China, Japan, Korea, and the surrounding seas (Wang, 2001; Wang et al., 2008). The mechanisms related to the EASM have been investigated extensively. Since the 2000s, the potential impact of AO on the EASM has drawn more attention.

The variability of the EASM-related climate is significantly correlated with the preceding spring AO (Fig. 7). Based on the summer rainfall data from six stations located over Yangtze River valley (east of 100°E) with record available from 1880 to 1999, Gong and Ho (2002) found that summer rainfall over Yangtze River valley is negative correlated with the May AO index with a correlation coefficient of -0.39 during 1899–1999, significant at 99% confidence level. They further revealed that an anomalous easterly around 30°–35°N and a stronger zonal wind along 40°–50°N in the East Asia are closely related to the positive May AO index, which is consistent very well with the relationship between summer monsoon rainfall and the East Asian jet stream. It implies that a stronger May AO is associated with a northwards movement of the summer jet stream, and a strong easterly as well as significant descending motion around 30°N. This gives rise to a drier condition in Yangtze River valley and a wetter anomaly over the western Pacific (Gong and Ho, 2003). The significant positive correlation between March–May AO and EASM (north of ~25°N) was further investigated by Gong et al. (2011) with focus on the importance of the North Pacific atmospheric circulation and SST. During March–May, the AO-associated atmospheric circulation anomalies produce warmer SST anomalies between 150°E–180° near the equator. The anomalous sensible and latent heating, in turn, induces an anomalous cyclone in the western North Pacific through a Gill-type response of the atmosphere. Through this positive feedback, the tropical atmosphere and SST patterns sustain their strength from spring to summer and finally weakens the summer western North Pacific subtropical high. Consequently, positive precipitation anomalies are developed over a broad region south of 30°N stretching from southern China to the western Pacific and the negative precipitation anomalies appear in the lower valley of the Yangtze River and southern Japan (As illustrated by Fig. 7).

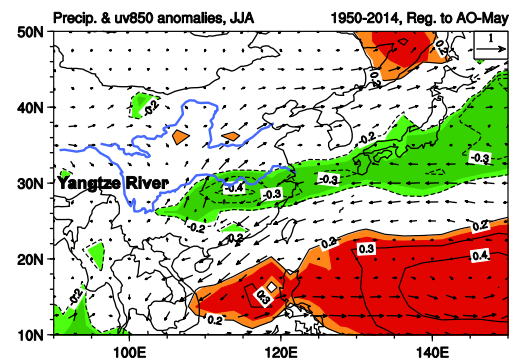


Fig. 7. Correlation coefficients (contours) of summer (June–August) precipitation and regression of 850-hPa wind (vectors) with the preceding May AO during 1950–2014. Shaded values indicate the correlation coefficients statistically significant at 95% confidence level, as estimated using the Student's *t*-test. Precipitation anomaly is provided by the NOAA/OAR/ESRL PSD, Boulder, Colorado, USA, from their Web site at <http://www.esrl.noaa.gov/psd/>. Wind datasets are derived from NCEP/NCAR Reanalysis. The figure is drawn on the method of Gong and Ho (2003).

However, the significant connection revealed by Gong et al. (2011) between the spring AO and the EASM might be unstable (Gao et al., 2014). Gao et al. (2014) found that the relationship between March–May AO and EASM on inter-annual timescale shows a remarkable decadal variation in late 1990s, with their correlation coefficient changing from +0.77 during 1979–1997 to −0.62 during 1998–2007. Following a spring positive AO phase, the lower-troposphere over East Asia features a cyclonic anomaly in summer before 1997 but an anticyclonic anomaly after 1997, which results from different simultaneous air–sea features imposed by spring AO and distinctive evolution from spring to summer. In pre-1997, the spring AO-associated wave activity prefers the high-latitude propagation from North Atlantic to Pacific, leading to the spring AO-associated signal being memorized and persisting over the Pacific. In contrast, a subtropical wave train from North Atlantic Ocean to Indian Ocean is evidently enhanced in post-1997 epoch, and accordingly the Indian Ocean plays a dominant role in memorizing and extending the influence of spring AO on EASM. Chen et al. (2015a) also documented an interdecadal change in the relationship between spring AO and the EASM which occurred around the early 1970. That is, the impact of spring AO on the following EASM is weak and insignificant during the 1950s and 1960s but strengthened and significant from the mid-1970s to the mid-1990s (Fig. 8), which might be caused by the interdecadal change in the intensity of spring North Pacific synoptic-scale eddy activity. Similar results were revealed by Li et al. (2008). They pointed out that positive phase of AO in March could lead to anomalous divergence in the lower troposphere where the Mei-Yu front is located, which is unfavorable for the upward motion. As a result, there is decreasing of Mei-Yu precipitation in the Yangtze River valley. They further suggested that the linkage between the March AO and Mei-Yu precipitation is unstable during 20th century. Significant negative correlation appears after the 1960s while significant positive correlation exists during 1940s and 1950s. Such a decadal change in the AO-precipitation relationship should be taken into account in the prediction of Mei-Yu precipitation (Gu et al., 2009). This might be one of the main reasons why Li and Leung (2013) drew a conclusion that relationship of the summer precipitation in China with AO from January to August is barely significant during 1960–2008.

Ju et al. (2005) suggested that the trend in the winter AO toward its high-index polarity after the late 1970s might promote the summer rainfall anomaly to changing from below normal to above normal in central China, the southern part of Northeast China and the Korean peninsula around 1978. It is revealed that the high-index polarity of winter AO leads to warming at mid- and high-latitude regions of the Asia continent, cooling at low-latitude regions in winter and spring and more precipitation over the Tibetan Plateau and South China. Due to the climatic memory, the increasing soil moisture makes the cooling tendency of the land in the southern part of Asia persist until summer. So the warming of the Asian continent is relatively slow in summer, causing an interdecadal decrease of the land-sea heat contrast and finally

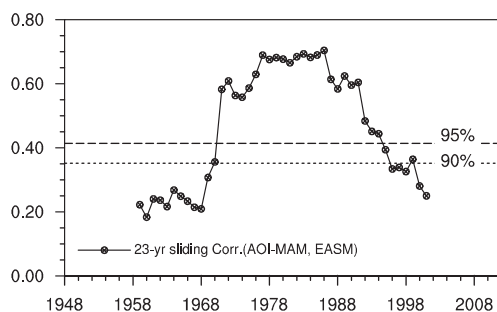


Fig. 8. 23-year sliding correlations between the spring (March–May) AO index and the EASM index. Results are obtained from NCEP/NCAR reanalysis datasets. Horizontal lines indicate the correlations statistically significant at 90%, 95% confidence level, respectively. Figure is repeated from Chen et al. (2015a).

leading to the weakening of the EASM circulation. Based on 600-year pre-industrial simulation by Bergen Climate Model Version 2, Cui et al. (2013) investigated the linkage between winter AO and the Southeast Asian summer monsoon (SEASM) on the inter-decadal timescale and found an in-phase relationship between the AO and SEASM with periods of approximately 16–32 and 60–80 years. The regression pattern between summer precipitation and winter AO for all 600 years showed a clear tri-pole distribution with significant positive rainfall anomalies over South China, the South China Sea and the Philippines. Cui et al. (2013) suggested that the large-scale anticyclonic anomaly in the subtropical North Pacific associated with the positive winter AO could cause northward warm (southward cold) ocean surface currents to the western-central North Pacific (northeast Pacific) regions, leading to the negative PDO. The AO-associated PDO-like winter SST anomalies could persist into summer and can therefore lead to inter-decadal variability of summer monsoon rainfall in East and Southeast Asia, as indicated by Deng et al. (2009). Otomi et al. (2013) also suggested that the AO polarity reversal from the record-breaking negative phase in 2009/2010 winter to a strongly positive one in 2010 summer might contributed to the record-breaking warm summer in Russia and Japan (In Japan, summer 2010 was the warmest in about 100 years of country-wide measurement records). Due to the influence of the negative AO in 2009/2010 winter, the winter SST in the North Atlantic Ocean showed a tri-polar anomaly pattern (warm SST anomalies over the tropics and high latitudes and cold SST anomalies over the mid-latitudes). The warm SST anomalies persisted into 2010 summer for the large oceanic heat capacity (which they called oceanic footprint), causing blocking highs to form over Europe which amplified the positive summertime AO and further strengthen warm conditions over Eurasia. Chen et al. (2014) suggested that in associated with spring positive AO, there is a pronounced anomalous anticyclone in the North Pacific and a pair of anomalous cyclones off-equator on both sides of the equatorial western Pacific, leading to anomalous westerly winds in the tropical western Pacific which could persist into summer through a Gill-like atmospheric response. The surface westerly wind anomalies in spring and summer propagate eastwards and enhance the tropical heating in summer, which in turn maintain the westerly wind anomalies persisting from summer to winter through positive feedback and further cause El Niño events. Similarly, Chen et al. (2013b) suggested that the positive spring AO could help the winter NPO to force an El Niño event via the seasonal footprint mechanism. However, when the spring AO is negative, the connection of the NPO and the El Niño is not robust at all.

East Asian summer climate is also related to the simultaneous AO. During 2009, the AO indices in June and July are the lowest during 1950–2009, and below-average air temperatures were also observed over North America and parts of Eurasia (L'Heureux et al., 2010). Tachibana et al. (2010) revealed that anomalies in positive summer AO accounts well for the hemispheric-scale weather associated with anomalous blocking between the polar and subtropical jets, whereas blocking rarely occurs during negative AO periods. The double jet stream structure is more evident during periods of anomalous positive AO than during periods of negative AO. The surface temperatures associated with the anomalous positive AO clearly show warm Europe and cool East Asia in summertime. During the development period (defined by a consecutive 11 day period starting from a day on which the AO index is less than + 1 standard deviation until a day on which the index is greater than + 3 standard deviation) of summer AO, blocking tends to occur over Europe and the Atlantic Ocean. On the other hand, the blocking tends to occur in the Far East Eurasia during the maintenance stage (period of 11 days during which the index continuously exceeds + 2 standard deviation). A significant connection between the tropical cyclone activities in the western North Pacific during July–August–September and the simultaneous AO has been found by Choi and Byun (2010). Compared to the positive (negative) AO years, more tropical cyclone activities formed over east (west) of 150°E, recurved

in the east (west), and passed over the midlatitudes (southeast Asian region), including Korea and Japan (South China Sea and south China) during negative (positive) AO years. This is because the anomalous anticyclone located around Korea and Japan during high-AO years could block the movement of tropical cyclones toward Korea and Japan. Instead, tropical cyclones moved westward toward southern China along the easterly and southeasterly steering flow of this anticyclonic circulation. As a result, tropical cyclone lifetime and intensity were shorter and weaker during high-AO years.

3.1. Summary of the mechanisms

The pathways for winter/spring AO impacting the East Asian summer climate can be summarized as follows: via changes in the soil moisture (e.g. Ju et al., 2005), SST changes in the North Atlantic (e.g. Cui et al., 2013; Otomi et al., 2013), the Indian (e.g. Gao et al., 2014) and the North Pacific (e.g. Gong et al., 2011) Oceans, and finally persists into the following summer. Through the above mechanisms, the East Asian summer climate shows a significant correlation with the preceding winter/spring AO. However, it should be noted that each pathway has only been investigated by limited literatures and no consensus has been reached. Long-term reliable observational data and coordinated modeling studies are needed.

4. East Asian spring dust storm, temperature, precipitation and AO

The frequently-occurring dust storms are another disastrous weather phenomena during spring (March–May) in North China (Fan and Wang, 2004, 2006, 2007). Besides, the variability of precipitation and temperature in spring is critical to food production in China. Therefore, the influence of AO on East Asian spring climate has been investigated in recent years.

The spring AO shows a significant out-of-phase correlation with the dust storm frequency in North China. Gong et al. (2006) found that the AO could modulate weather and dust storm activities by influencing local baroclinicity. During the positive-AO springs, there is a smaller eddy growth rate in northeastern China and its vicinities. As weather changes are closely related to the atmospheric instability, the synoptic variance tends to be weaker during the positive-AO springs. Consequently, it would give rise to less dust storm. Liu and Ding (2007) revealed that the spring atmospheric baroclinic structure might be related to the preceding winter positive AO, which is unfavorable for the occurrence of the spring dust storm frequency in Northwest China. The potential impact of April AO on the interannual variability of April dust emissions from the Tarim Basin was further investigated by Gao and Washington (2010) for the period from 1980 to 1992. They found that in the positive phase of AO, geopotential gradient between the inside and outside of the basin is small. The northerly wind from Siberia is weak and decreases when it encounters the north barrier of the basin. As a result, dust activities in the Tarim Basin are low. In the negative phase of AO, geopotential gradient between the inside and outside of the basin is large. The northerly wind from Siberia is strong, and passes the north barrier of the basin. When it encounters the Tibetan Plateau, a component of the wind goes into the basin. During this process, the wind accelerates due to the higher pressure gradient. As a result, dust activities in the Tarim Basin are frequent and intensive. With combination of atmosphere reanalysis data and model simulation, Mao et al. (2011b) investigated the northeast Asian dust process during 1982–2006 spring seasons and found that a positive AO phase is associated with decreased (increased) dust storm frequency in Mongolia (Taklimakan Desert) and enhanced anticyclonic (southeastward) dust transport over northwestern China (North China). The related mechanism is that a positive AO phase induces a northward shift of the polar jet and an intensified westerly jet over northern Tibetan Plateau. The former reduces the frequency of intense cyclones in Mongolia, thereby causing a decrease in the dust storm frequency. The latter increases

the dust storm frequency in the Taklimakan Desert. Further, the positive geopotential height anomaly over Mongolia in association with positive AO initiates an anti-cyclonic dust transport anomaly in the middle troposphere over northwestern China and induces a southeastward dust transport anomaly over North China. Based on the atmosphere reanalysis data, dust storm records, and the numerical simulations with the dust model, Mao et al. (2011a) further pointed out that there is a significant linkage between the spring dust storm frequency and AO; a negative (positive) AO phase is related to an increased (decreased) dust storm frequency in North China. They suggested that the AO has large influence on the frequency of cold air activity over Mongolia through its impact on the cold air activity. Recent study indicated that the ENSO might modulate the relationship between the spring dust activity over northern China and AO. The frequencies of spring dust events (e.g. dust storm, blowing dust, and floating dust) over northern China (35°N–45°N, 105°E–120°E) increase significantly in the years with negative AO and El Niño compared with those in the years of negative AO and La Niña (Lee and co-authors, 2014). The possible physical mechanism is that the anomalous large-scale environments (such as the intensified the zonal wind shear and atmospheric baroclinicity) associated with negative AO and El Niño are more effective to provide favorable conditions to enhance Asian dust activity (Lee and co-authors, 2014).

A positive winter AO polarity could lead to spring SAT rise over north Eurasia and SAT decreasing over Northwest China. The solar irradiance and surface state (the snow/ice-albedo) are important factors to explain the variance of SAT during spring (Kryjov, 2002). During the positive AO winters, due to enhanced advection of the warm Atlantic air, there are positive SAT anomalies over northwestern Eurasia, which leads to more fractured ice and spring snow/ice melt and further causes lower albedo and spring SAT rise. Through such a positive feedback, the influence of the winter AO on spring SAT is evident at least until the end of snow/ice season. The winter AO accounts for some 25–50% (15–20%) of the simultaneous (spring) SAT variance in northern Russia (Kryjov, 2002). The mechanism proposed by Kryjov (2002) is supported by the results revealed by Bamzai (2003) that the AO index during January–March shows a significant inverse relationship with snow cover in concurrent and subsequent spring months, particularly over Eurasia. Besides, the hypothesis that increased temperatures and decreased precipitation due to a positive trend in the winter AO have advanced the time of January–March snow melting in the northern hemisphere was tested by both reanalysis datasets and model simulation (Schaefer et al., 2004). However, on the subseasonal time scale, the influence of the winter AO on the monthly temperature fields over East Asia varies. Huang et al. (2007) revealed that deeply influenced periods are in May, June and August. The most significantly influenced region in spring is located in Northeast China. There are significant positive correlations between the previous winter AO and spring SAT over Northeast China (Fig. 9). Besides, Liu and Ding (2007) suggested that there are significant negative correlations between the previous winter AO and the spring temperature over Northwest China as well as significant positive correlations between the previous winter AO and the spring 500 hPa geopotential height over the Mongolian Plateau and Middle Siberia. A significantly negative relationship between the December–March AO and the following spring extreme low temperature events in the north of eastern China was also revealed by Yin et al. (2013). The possible physical mechanism is that, in associated with positive (negative) winter AO, the mid-latitudes of Eurasia is warmer (colder), leading to smaller (larger) than normal Eurasian snow cover area in winter which could strongly persist to the following spring. Accordingly, spring cold vortices over Northeast China are weaker (stronger) and the spring SAT in the north of eastern China is higher (lower), resulting in less (more) extreme low temperature events. Additionally, the relationship between the winter AO and surface-climate anomalies in the following spring might be modulated by the 11-year solar cycle (Chen and Zhou, 2012). The spring temperature anomalies in northern China related to the previous winter AO were larger and more robust after high solar

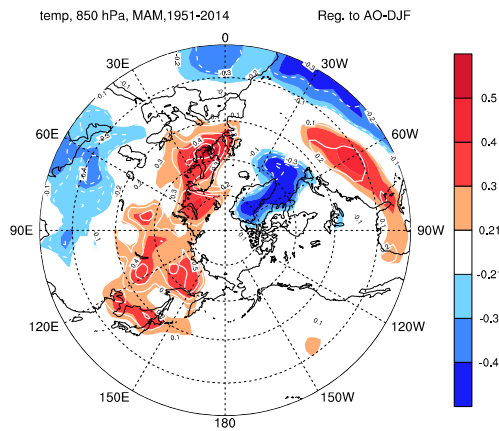


Fig. 9. Regression of spring 850-hPa air temperature (°C) derived from NCEP/NCAR reanalysis dataset upon the preceding winter AO index during 1950–2013. Shaded regions indicate the correlations that are statistically significant at 90% confidence level, as estimated using the Student's *t*-test. The figure is drawn on the method of Liu and Ding (2007).

cycle winters. However, spring temperature anomalies became very small and insignificant after the low solar cycle winters.

The potential impact of winter/spring AO on the East Asian spring precipitation has not drawn much attention. The correlation between the previous winter AO and the spring precipitation in Northwest China isn't significant on an interannual time scale, whereas there is a positive correlation area in Northwest China between the previous winter AO and the spring precipitation on a decadal time scale with correlation coefficients significant at the 95% confidence level (Liu and Ding, 2007). A negative correlation between the rainfall over North Korea in May and the simultaneous AO is revealed by Choi et al. (2013). When the May AO is in positive phased, anomalous anticyclone is strengthened in the Maritime Province of Siberia, whereas anomalous cyclones have relatively been reinforced in the sea east of Japan. Therefore, as North Korea was affected by the cold and dry anomalous northeasterlies derived from the anomalous anticyclone located in the Maritime Province of Siberia, North Korea's May total rainfall resulted in a decline. Besides, the anomalous cyclone strengthened in the sea east of Japan indicated that the western North Pacific subtropical high in the positive AO phase was less developed than in the negative AO phase, leading to decreasing of warm and humid air to North Korea. In addition, the anomalous cold SSTs were strengthened in the seas of the mid-latitude East Asian in the positive AO phase, thereby causing atmospheric stabilization, and this ultimately contributed to the decline in North Korea's May rainfall.

4.1. Summary of the mechanisms

AO could modulate the variations of spring dust storm frequency over East Asia through its impact on the local baroclinicity (e.g. Gong et al., 2006; Liu and Ding, 2007; Lee and co-authors, 2014), the geopotential gradient (e.g. Gao and Washington, 2010), the shift of polar jet and the intensity of westerly jet over northern Tibetan Plateau (e.g. Mao et al., 2011a, 2011b), and the cold air activity (e.g. Mao et al., 2011a). Positive winter AO polarity generally causes positive (negative) spring SAT anomalies over Mongolia and Northeast China (Northwest China) through a snow/sea ice–albedo positive feedback mechanism (e.g. Kryjov, 2002; Liu and Ding, 2007).

5. Possible climate predictability using Arctic oscillation

The anomalous AO could impact the subsequent East Asian climate and is usually preceded by significant anomalous signal, especially on seasonal to inter-seasonal time scales. Wang and Chen (2010) revealed

that the extreme cold December 2009 in Eurasia is closely related to the extreme negative AO, which is the lowest one during 1979–2009. They pointed out that the negative AO in December 2009 is closely associated with the downward propagation anomalies from the stratosphere which are associated with the strong negative November AO in the stratosphere. In November, the stationary planetary wave propagation along the polar waveguide into the stratosphere is anomalously strong. In December, there exists anomalous downward propagation of stationary planetary wave around 50°N from the stratosphere to the troposphere, which induces a robust negative AO in December. Their results imply a potential prediction of winter climate (e.g. mean-state of temperature, cold surges, and cold waves) by the preceding anomalous AO in sub-seasonal time scale. It sounds reasonable because the downward propagation of AO signal from the stratosphere is usually accompanied with preceding upward and then downward propagation of planetary wave anomalies (Christiansen, 2001; Cohen et al., 2010). For example, Kim and Ahn (2012) investigated the relationship between the East Asian winter temperature and the preceding AO during 1978–2007 and found that the AO index in November has the highest positive correlation coefficient with the winter SAT. Their analysis further indicated that following a positive (negative) phase of November AO, the Siberian high weakens (strengthens), which in turn weakens (strengthens) the EAWM, resulting in a warm (cold) winter in East Asia. Their results implied that the November AO might be used in forecasting the winter climate over East Asia. However, He and Wang (2013b) found that the correlation of the November AO with the following January East Asian temperature is more significant than that with the following December and February for the period of 1950/51–2010/2011. Moreover, the January temperature in East Asia is also closely related to the preceding December AO. The related mechanisms are as follows. A Rossby wave associated with the November AO is confined to high latitudes in December, but shifts southeastward to East Asia in January. Similarly, the December AO-related wave activities propagating southeastward to East Asia could persist into the following January. Consequently, the signals of the November/December AO could be transmitted to the following January. Besides, an air–sea interaction might exist over the North Pacific. The SST over the central subtropical North Pacific (west coast of North America) often rapidly rises (drops) a month later and peaks in the following January when the preceding November/December AO is in positive phase. Through the above processes, a positive phase of November/December AO is generally followed by weaker-than-normal Siberian high, Aleutian Low, East Asian trough, and a weakening of meridional shear of East Asian jet in January (Fig. 10), which would favor warmer conditions in East Asia. What's more, the influence of the November AO could even persist into the following spring and summer. Through an interaction between the synoptic-scale eddy and the low-frequency mean flow over the subtropical North Pacific, the positive (negative) phase of November AO tends to induce positive (negative) SST anomalies in the tropical central-eastern Pacific during the following spring and summer (Chen et al., 2015b). It implies that the November AO index might be considered as a potential predictor of ENSO-related SST anomalies during the following spring and summer and further for the East Asian climate. Besides, due to the re-emergence of SST anomalies over the western North Atlantic, positive phase of January–February AO tends to cause positive SLP anomalies occupying the entire Northern Eurasia and negative SLP anomalies spanning from the Bering Sea to the Western North Atlantic in the following November (Choi et al., 2016). It is also revealed that the extreme cold (warm) events are less (more) frequent in January over East Asia when the preceding (October–December) AO is anomalous higher than normal, which might be attributed to the intrinsic persistence of AO in stratosphere and the memory of Eurasian snow cover (He and Wang, 2016). However, it should be noted that the frequency of January cold events might show a more significant relationship with the negative phase of AO than that with the positive phase of AO (He, 2015b).

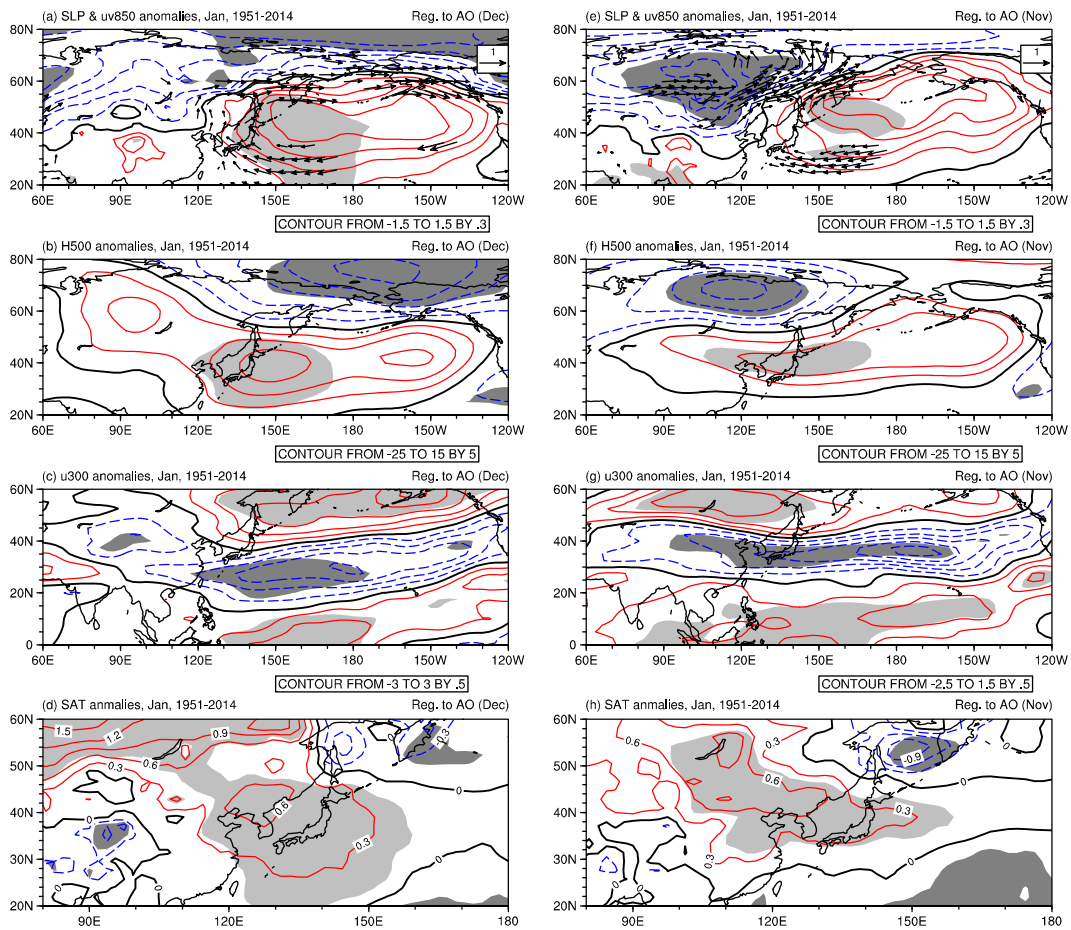


Fig. 10. Regression maps of the following January (a) 850-hPa wind (vector), sea level pressure (shaded), (b) 500-hPa geopotential height, (c) 300-hPa zonal wind, and (d) surface air temperature anomalies onto the preceding December AO index during 1950–2013. (e)–(h) are the same as (a)–(d), but for the November AO index. The vector winds and shaded region are significant at 95% confidence level, based on a Student's *t*-test. The figure is drawn on the method of He and Wang (2013b).

It is also found that following a positive phase of spring AO, positive summer precipitation anomalies are observed over southern China to the western Pacific and the negative precipitation anomalies appear in the lower valley of the Yangtze River and southern Japan (Gong and Ho, 2003; Gong et al., 2011). Besides, higher (lower) surface temperatures generally appeared in Mongolia and Northeast China (Northwest China) during spring after a positive AO winter, and vice versa (Liu and Ding, 2007; Chen and Zhou, 2012). Therefore, the climate prediction based on the previous AO is worth further study.

6. Summary and discussion

A most dominant feature of atmospheric circulation is the atmospheric teleconnection patterns (also refers to Oscillation, seesaw patterns), which are made up of chains of lobes whose centers are connected by lines that have prominent meridional components (Branstator, 2002). During the 20th century, many teleconnection patterns have been identified. Among these teleconnection patterns, the NAO has drawn the most attention because of its significant climatic impact. By the end of 20th century, a new teleconnection (i.e. AO) has been found by Thompson and Wallace (1998). Numerous atmospheric scientists have documented that the AO could impact significantly the climate over Europe and Far East. Considering the fact that the AO and NAO have similar climatic impact over Europe and there has been no review paper on AO, in this paper, we have reviewed the impact of AO with a special focus on the region of East Asia.

This review synthesizes the published literature concerning the impact of AO on the EAWM (related cold surges/cold wave), East Asian

winter precipitation, EASM, East Asian summer precipitation and spring dust storm, temperature, and precipitation.

It is evident that a positive winter AO causes warmer winters over East Asia through enhancing Polar westerly jet which prevents cold Arctic air from invading low latitudes (Thompson and Wallace, 2000). Many authors also conclude that a positive winter AO could lead to a weaker-than-normal EAWM and warm conditions over East Asia through its impacts on the Siberian high (Gong et al., 2001; Wu and Wang, 2002a), Rossby wave (Park et al., 2011), and the polar and sub-tropical jets (Li et al., 2014). The opposite is true concurrent with negative AO phases. By contrast, some authors assert that the winter AO influences directly the winter SAT and EAWM-related atmospheric circulation over the region northwards of 35°N (Wu and Wang, 2002b). Therefore, there remains a major challenge to understand the pathway that the AO impacts the EAWM.

Meanwhile, the effect of winter AO on the extreme climate has also been extensively studies. Results drawn from several studies applying re-analysis datasets and observational datasets have yielded the same conclusions that negative phase of AO leads to more frequent cold air outbreaks over East Asia, and vice versa (Thompson and Wallace, 2001; Chen et al., 2005; Jeong and Ho, 2005; Wei and Lin, 2009; Wen et al., 2009, 2013; Park et al., 2010). Cold air outbreaks can take place under positive and negative phases of AO, but with different pathways (He, 2015b). The cold surges during a negative AO were modulated by southward expansion of the Siberian high, which originated in the Arctic regions. On the other hand, the cold surge during a positive AO occurred when the Siberian high extended southeastward across the Eurasian continent (Park et al., 2010). Hence, in any of the coming

winters with anomalous AO patterns, understanding the development of the Siberian high is still an important way to improve the possibilities for seasonal forecasting of cold surges/waves over East Asia.

Although most of the recent work has addressed the dominant effects of AO on the winter climate over East Asia, some studies indicate that the impact of AO on East Asian winter climate is spatially and temporarily dependent (Woo et al., 2012; Chen et al., 2013a; Li et al., 2014) and might be modulated by external forcing (Quadrelli and Wallace, 2002; Chen and Zhou, 2012; Ando et al., 2015). Thus, it is by no means guaranteed that the anomalous weather and climate which the AO tends to generate will dominate in East Asia. We should be cautious in the climate prediction based on AO.

It is found that a positive spring AO is usually followed by significant positive summer precipitation anomalies in southern China and western Pacific as well as negative precipitation anomalies in the lower reaches of Yangtze River and southern Japan (Gong and Ho, 2003; Gong et al., 2011). Whereas, the mechanisms behind them are still unclear. What's more, the spring AO-EASM relationship is unstable (Gao et al., 2014; Chen et al., 2015a).

Many recent studies suggest that the spring AO shows a significant out-of-phase correlation with the interannual variations of the dust storm frequency in northern China, through its influence on the local baroclinicity (e.g. Gong et al., 2006; Liu and Ding, 2007; Lee and co-authors, 2014), the geopotential gradient (e.g. Gao and Washington, 2010), the shift of polar jet and the intensity of westerly jet over northern Tibetan Plateau (e.g. Mao et al., 2011a, 2011b) and the cold air activity (e.g. Mao et al., 2011a).

Even though a robust negative AO-EAWM relationship has been generally accepted, what causes such a connection is still an open question. Gong et al. (2001) suggested that the winter AO could influence the EAWM through the impact on the Siberian high. Wu and Wang (2002a) also pointed out that the winter AO could influence the winter Siberian high through its impact on the downward motion of the nearly entire troposphere. However, Wu and Wang (2002b) further asserted that the winter AO influences directly the winter SAT, SLP, and the East Asian trough (500 hPa) over the region northwards of 35°N in East Asia rather than through its impact on the Siberian high. Nevertheless, Gong and Wang (2003) demonstrated the correlation coefficient between the AO and the average temperature in China during 1958/59–1994/95 winters is 0.43, which decreases dramatically to 0.14 when the contribution of the Siberian high is excluded.

Additionally, most of the previous studies in East Asian winter climate focused on the seasonal mean such as three-month mean (i.e. December, January, and February or January, February, and March) or five-month mean (i.e. November, December, January, February, and March). However, some recent observations indicated that the East Asian winter climate between different sub-seasonal time scales such as early winter and late winter shows significant out-of-phase relationship (Chang and Lu, 2012), suggesting the winter mean might cover up some primary fact in scientific research and application. For example, Kim and Ahn (2012) found a close connection between the November AO and the East Asian winter temperature during 1978–2007. However, He and Wang (2013b) found that the correlation of the November AO with the following January East Asian temperature is more significant than that with the following December and February for the period of 1950/51–2010/2011. Moreover, the January temperature in East Asia is also closely related to the preceding December AO. It implies that a large intraseasonal variation might exist in the relationship between the AO and East Asian winter SAT anomalies. For example, a weak in-phase relationship of AO with the southern China SAT anomalies occurs in December, but a significant out-of-phase relationship occurs in January and February (Zuo et al., 2015). Therefore, further studies should pay more attention to the impact of AO on the East Asian climate on the sub-seasonal time scale. The negative trend of winter AO during 1990–2013 could describe the Eurasian cooling, but failed to explain the Arctic warming (Cohen et al., 2014). The Ural blocking might

provide a new insight in exploring the mechanisms for the formation of warm Arctic-cold Eurasia (Luo et al., 2016a, 2016b), which should be paid more attention in further studies. Besides, as the results in phases 3 and 5 of the Coupled Model Intercomparison Project (please see <http://cmip-pcmdi.llnl.gov/>) are released, many studies discussed the climate projection over East Asia (Hu et al., 2000; He and Wang, 2013a, 2013b; Jiang et al., 2013; Jiang and Tian, 2013). However, the projection of AO and its impact on the East Asian climate in future is barely explored.

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